

## Observation of a magnetic sub-storm event in the equatorial thermosphere-suggestion for interpretation

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**Abstract** The equatorial peak atomic oxygen density in the lower thermosphere (around 100 km) is seen to be moderately enhanced following a magnetic sub-storm using data obtained from the ground based observations. The temperature of the mesopause obtained from OH(7 - 2) night airglow observation has shown a rise similar to the positive bay of the magnetic field intensity (horizontal component) during the said disturbance. There is no enhancement of OI(6300) intensity around 300 km compared to the OI(5577) rise. The implication of these observation is discussed in the light of the extra-ionospheric current, increasing the recombination rate.

**Keywords** Thermosphere, magnetic sub-storm, airglow

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### 1. Introduction

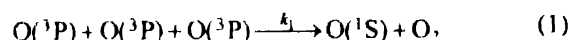
Variations in the total density of the thermosphere during magnetic disturbances have been first reported as early as in 1959 at different altitudes and latitudes [1]. Subsequently, extensive information on the same was obtained using the satellite-borne gas analyzer [2]. Evidence for changes in OI(6300) intensity have also been observed during the magnetic storms [3,4] which suggest variations in neutral oxygen density. Pelz *et al* [5] have measured the densities of oxygen during a magnetic storm at altitudes down to 160 km by using data from AE-C satellite. Information pertaining to the behaviour of neutral oxygen density during the same activity at the lower thermosphere is of great importance owing to the negative ion chemistry of the D-region. Weill and Christopher-Glaume [3] have observed increase in the intensities of OI(5577), O<sub>2</sub> band and continuum and decrease in OH intensities at mid latitude station during and following magnetic storms. Contrary to this, Shefov [6] found an increase in OH intensities and a reduction in the rotational temperature following the storms.

Following a ground based observation made at Kodaikanal (10°N, 77°5'E), we found that there is a temporal enhancement of atomic oxygen at an altitude of 100 km density soon after

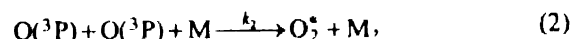
a magnetic sub-storm. The mesopause temperature is also determined from OH(7 - 2) lines. Finally, comparison of the behaviour of atomic oxygen lines OI(5577) and OI(6300) is made.

### 2. Method for deducing atomic oxygen density around 100 km

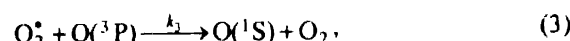
The most widely accepted reaction rates for the production of OI(5577) radiation are those suggested by Chapman [7], viz.,



and by Barth [8],



followed by,



$k_1$ ,  $k_2$ ,  $k_3$  being the respective reaction rates. We follow the reaction suggested by Chapman [7] in the present study.

The volume emission rate of the green line of atomic oxygen due to the Chapman reaction is given by

$$Q = \frac{k_1(\text{O})^3}{1 + [k_4(\text{O}) + k_5(\text{O}_2)]/A}, \quad (4)$$

$k_4$  and  $k_5$ , being the respective quenching coefficients for the reactions  $O(^1S) + O(^3P) \xrightarrow{k_4} 2O(^1D \text{ or } ^3P)$  and  $O(^1S) + O_2 \xrightarrow{k_5} O + O_2$ .  $A$  is the Einstein transition probability coefficient from  $O(^1S)$  to  $O(^1D)$ . For calculating the column emission rates of the green line, the following values for  $k_1$  [9],  $k_4$  [10]  $k_5$  [11] and  $A$  [12] are used :

$$k_1 = 1.4 \times 10^{10} \exp[-650/T] \text{ cm}^6 \text{ mol}^{-2} \text{ s}^{-1}, \quad (5)$$

$$k_4 = 5.0 \times 10^{11} \exp[-305/T] \text{ cm}^3 \text{ mol}^{-1} \text{ s}^{-1} \quad (6)$$

$$k_5 = 4.0 \times 10^{12} \exp[-865/T] \text{ cm}^3 \text{ mol}^{-1} \text{ s}^{-1} \quad (7)$$

$$A = 1.17 \text{ s}^{-1}. \quad (8)$$

It is evident from the neutral atmospheric model [13] that the quenching due to  $N_2$  is insignificant. Therefore, the expression for column emission rates can be expressed as :

$$I = \int_0^\infty \frac{k_1(O)^3 dh}{1 + [k_4(O) + k_5(O_2)]/A}, \quad (9)$$

the integration being done over the emitting region.

Reed and Chandra [14] have evaluated the intensity  $I$  numerically as follows :

$$I_{\lambda(5577)} = \frac{0.216 k_1 S H (O)_m^3}{1 + [0.8 k_4 (O)_m + k_5 (O_2)_m]/A} \quad (10)$$

in which  $S$  denotes the mixing scale factor.  $H$  is the scale height of mixed atmosphere expressed as  $H = kT/mg$ , where,  $k$  is the Boltzmann constant,  $g$  the acceleration due to gravity at peak neutral oxygen density  $(O)_m$ ,  $m$  the mean molecular mass  $\approx m(N_2)$ .  $T$  is the absolute neutral gas temperature near the peak oxygen density. Eq. (6) is used to deduce the maximum oxygen density  $(O)_m$  in the lower thermosphere. Rao and Murthy [15] computed the variations of  $(O)/(O_2)$  with temperature for different values of  $S$  at various heights between 95 and 100 km using the model  $(O)$  profiles obtained by them and the values of  $(O_2)$  densities taken from CIRA [13]. Since the height of maximum  $(O)$  density  $Z_m$  is directly proportional to the product of  $S$  and  $H$ , the altitude of  $(O)$  maximum varies with various values of  $T$  and  $S$ . The height of  $Z_m$  is calculated using the relation  $Z_m = 76.9 + 4.9 (S.H)$  with observed value of temperature. Therefore, the ratio of  $(O)/(O_2)$  is taken from the model computed by Rao and Murthy [15] at various temperatures obtained observationally and at different heights of  $(O)$ , keeping the values of  $S = 0.7$ , since the parameter  $S$  cannot be obtained from our observational data.

The temperature at the lower thermosphere region is determined from OH(7-2) night airglow observations. The ratio of intensities of  $P_1(3)$  line (6922.8) and  $\Sigma R$  branch (6838) of OH(7-2) band is measured experimentally. The same ratio is also calculated theoretically by using the

following equation at different temperatures from synthetic spectrum of OH(7-2) band :

$$I' = C s_j \exp[-F(J)hc/kT], \quad (11)$$

where  $I'$  is the intensity of rotational line,  $s_j$  the line strength,  $J$  the upper level rotational quantum number,  $T$  the rotational temperature,  $C$  an arbitrary constant. The function  $F(J)$  is defined by  $B_v J(J+1)$  where  $B_v$  is the effective rotational constant. The rotational temperature of OH is considered to be the true temperature of the atmospheric level where OH is dominant. Therefore, the temperature is determined by comparing the ratio obtained experimentally with that obtained theoretically using eq. (11).

It is generally agreed that the maximum of OH density occurs at 85-90 km and so the maximum emission in OI(5577) which is 10 km above the dominant OH emission can be determined. The steady temperature profile [13] shows that the ratio of temperatures at the 100 km and 90 km levels is about 1.115. In the first approximation, we can assume that the temperatures at two levels are always related by such a factor. Since the factor is based on a mean model, it cannot be taken as constant for determining the variations in temperature ratio at different locations and times. The derived number density of neutral oxygen is therefore subject to the uncertainties in the value of this temperature ratio.

It is to be noted that along with  $E$ -region component, the observed intensity of OI(5577) includes the contribution from the  $F$ -region as well, arising from the dissociative recombination of  $O_2^+$ . Therefore, to estimate the  $E$ -region component of OI(6677), 25% of the OI(6300) intensity measured simultaneously has been subtracted from the observed OI(5577) [16,17].

### 3. Observations and results

A filter photometer was used to monitor the intensities of OI(5577), OI(6300), OH(7-2) bands  $\Sigma R(6838)$  and  $P_1(3)$  line of OH(6922.8) from the zenith sky together with the background continuum at 7120Å. The half-power bandwidths of interference filters used are 10.03Å, 11.50Å, 22.53Å, 8.07Å and 40Å, respectively. The calibration is done with  $^{14}C$  source. The filter wheel takes 12 minutes to complete a rotation.

Figure 1 depicts the intensities of OI(5577), OI(6300) along with rotational temperature obtained from OH(7-2) night airglow observations on the disturbed night namely, 28-29th April, 1984. The error is found to be about  $\pm 3\%$ . The same quantities for the quiet night 27-28th April, 1984 as well as 2-3rd May, 1984, the first available post sub-storm night are presented in Figures 2(a) and 2(b) respectively. The height of the  $F$ -layer ( $h'F$ ) records for the respective nights

obtained from the ionosonde working in the same location are also plotted in the afore-mentioned figures at the interval

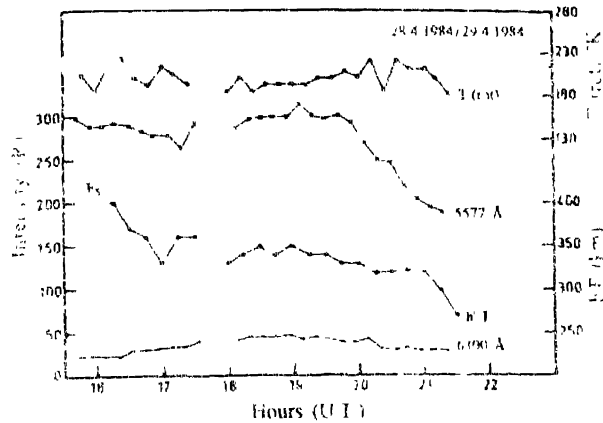


Figure 1. The intensities of OI(5577), OI(6300) along with the height of  $F$ -layer ( $h'F$ ) ionization during the disturbed night 28–29 April 1984. The curve at the top shows the rotational temperature obtained from the (17, 2) band.

of 15 minutes. The gradual increase in the intensity of OI(6300) during midnight on the quiet days, viz., 27–28th April as well as 2–3rd May (see Figures 2a and 2b) is

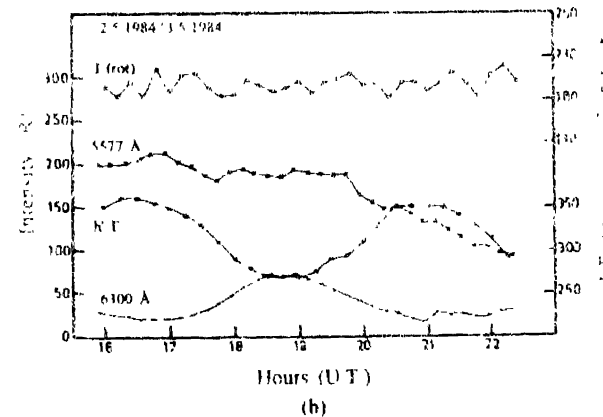
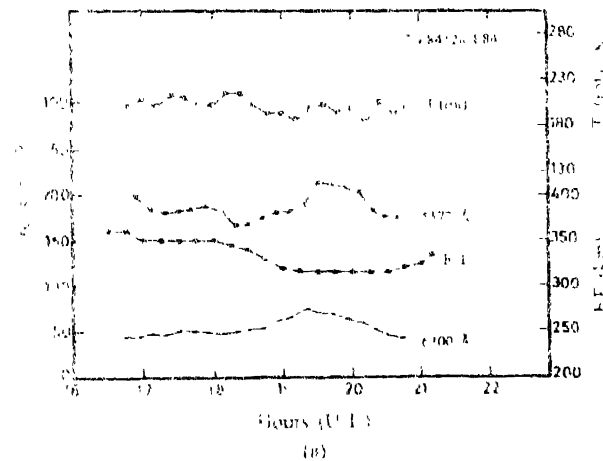


Figure 2. (a) The intensities of OI(5577), OI(6300) and the temperature along with the height of the  $F$ -layer ( $h'F$ ) ionization on the pre-substorm night, 27–28 April, 1984 and (b) during a quiet night, 2–3rd May, 1984.

primarily attributed to the downward movement of ionization in the  $F$ -region [18]. In the disturbed night, the downward movement of the  $F$ -region ionization was sharp during the decay phase of the sub-storm but it remained stable around 340 km with minor variations. As a consequence, the intensity of OI(6300) did not rise significantly throughout the disturbed night. It is to be noted here that the ionograms obtained from the ionosonde showed the spread- $F$  during and following the sub-storm which lasted till midnight, but the disturbance recorded in the magnetic variometer, shown in Figure 3, indicates an unusual smooth variation after the recovery phase. The three-hourly  $K_p$  value was 4 during which the sub-storm was recorded and it was 2<sup>+</sup>, 2<sup>+</sup> for the subsequent intervals following the sub-storm. The  $A_p$  value was 17 on the said night.

The lower part of Figure 3 depicts the temporal changes of oxygen density in the lower thermosphere during the

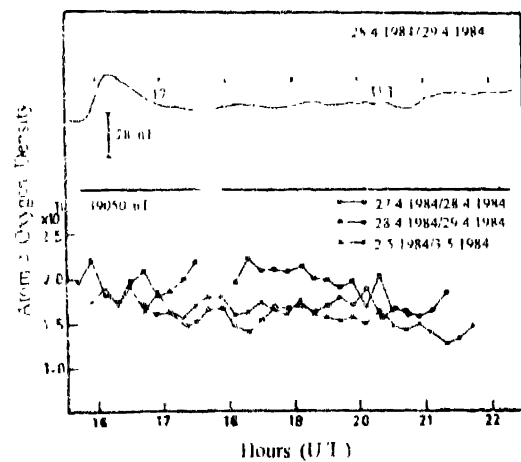


Figure 3. The temporal changes of peak atomic oxygen density in the lower thermosphere (around 100 km) during the disturbed night (28–29 April, 1984), as well as during the quiet nights (27–28th April, 1984), the pre-storm night and 2–3rd May, 1984, the first available post-substorm night. The upper panel shows the horizontal component of the earth's magnetic field of the disturbed night.

disturbed night, 28–29th April, 1984, together with measurement taken during the quiet nights, 27–28th April, 1984, as well as on 2–3rd May, 1984, which serve as a reference. Following the sub-storm, the atomic oxygen density shows a moderate enhancement of the order of ~23% from the quiet night values. In contrast, the variations of the said density during the control nights were found to be of the order of ~5–6%. The three-hourly  $K_p$  values were respectively 3, 3, 4 and 2<sup>+</sup>, 1, 1 for the pre, as well as for post sub-storm night. The  $A_p$  values are 27 and 17, for the respective nights. The upper panel, in Figure 3, shows the intensity variations of the horizontal component of the earth's magnetic field recorded at the same station. The observed rotational temperature (see Figure 1), shows a

similar hump as the bay in the magnetic records at the time of growth and decay phase of the sub-storm, with a time lag of 15 minutes. However, the temperature did not change much during the rest of the period as compared to the quiet nights. The gap in Figure 1 is due to power failure which lasted about 20 minutes.

#### 4. Discussions

##### 4.1 Absence of $OI(6300)$ line emission

The data obtained on  $OI(6300)$  intensity at Kodaikanal did not show any enhancement as compared to the study of Misawa *et al* [4], during the disturbed night. Various investigators have tried to explain the changes in the neutral composition associated with the storm, arguing that (i) combination of wind-induced transport and thermal expansion would reproduce the composition at altitudes higher than 150 km [19–21] and (ii) transfer of energy from high to low latitudes by means of gravity waves as suggested by Testud [22]. However, Rishbeth *et al* [23] mentioned about the large time lag of transmission of the changed composition from the turbopause to the higher altitude. According to Prölss [2], traveling atmospheric disturbances may contribute to the early onset of the low latitude perturbation. Thus, the time delay between the peak of the disturbance and the associated effect on the atmosphere in an equatorial station may be the cause of the absence of  $OI(6300)$  peak as exhibited in the present study.

Jacchia *et al* [24] found that the mean time delay between the peak of the geomagnetic disturbance and its effect on the atmospheric composition for the mid latitudes is of the order of 6 to 7 hours. On the other hand, Prölss [25] mentioned the 4 hours delay of response time of the low latitude upper atmosphere to the high latitude magnetic activity. Consequently, the hump of equatorial density would be discernible only after the transport time of latter which is of the order of 6 hours well after our observing schedule (limited by the moonlight). This conclusion is substantiated by the fact that, using Cactus accelerometer data, Berger and Barlier [26] found a time lag ranging from 4.6 to 6 hours for the response of the enhancement of total density in the equatorial zone with respect to  $AE$  index for the early afternoon sector and  $\sim 4.6$  hours for the midnight sector.

Thus far, we have discussed the absence of intense peak of the red line following the sub-storm. However, we shall now look at another phenomena that occurred during the disturbed night. This concerns the temporal variation of  $OI(6300)$  intensity during the course of a night. It is well established that the temporal variations of the  $OI(6300)$  intensity in the equatorial zone depend on the corresponding changes of vertical  $E \times B$  plasma drift velocity. Owing to the reversal of the electric field after the sunset, an equator-

ward movement of the ionization ensues in the Appleton anomaly region. This, in turn, produces a downward movement of the ionization over the equatorial zone giving rise to the increase of the recombination rate. Consequently, the intensity of  $OI(6300)$  emission would seem to be enhanced [18, 27–30]. This phenomena has been observed on the quiet nights in the present study of  $OI(6300)$  intensity accompanied by a downward movement of the  $F$ -region ionization. However, in the disturbed night, the intensity of  $OI(6300)$  seemed to be constant. An inspection of the ionograms obtained from Kodaikanal (see Figure 1) shows that the height of the ionization in  $F$ -region remained steady following the storm. This confirms the similarity of behaviour between the ionization height and red line intensity as shown by various investigators [17,31].

##### 4.2 Presence of $OI(5577)$ line emission

In contrast to the above fact, the change in maximum oxygen density ( $O_m$ ) in the lower thermosphere after the recovery phase of the sub-storm is significant. In fact, the intensity of the  $OI(5577)$  has shown an enhancement of the order of 50% from the quiet night values which in turn, causes an increase of the population of  $O(^1S)$ , the metastable state transition from which gives wavelength  $OI(5577)$  in the lower thermosphere during, as well as following the sub-storm. It may be noted here that the effect due to the aforementioned disturbance sustained to a great extent of time at the lower thermosphere. The diffusion time  $\tau_0 \sim 4.5$  hrs at that height  $[-2\sigma vHN/g]$ , where  $\sigma$  is the collision cross section for atmospheric gases,  $v = \sqrt{3kT/m}$  the mean thermal speed,  $H$  the scale height,  $N$  the background gas density,  $g$  the acceleration due to gravity,  $k$  the Boltzmann constant,  $T$  the absolute temperature,  $m$  the mean molecular mass of the gas], could be the cause of this phenomena [32]. Weill and Christophe-Glaume [3] have also found marked increase in  $OI(5577)$  during and following the storm at mid latitude which was postulated by Thomas [33] as due to the increase in  $O$  concentration on the upper D-region. However, changes in the atmospheric composition in the tropopause height in connection with changes in the thermospheric temperature during the magnetic sub-storm have also been postulated [34,35]. In view of the immediate enhancement of atomic oxygen density following the sub-storm, our results cannot be explained by the transport of  $O$  atoms in the lower ionosphere from the auroral belt to the equator as suggested by Thomas [33]. On the other hand, the extra-ionospheric currents [36] which turns out as the positive bay (see Figure 3), may be responsible for heating the lower ionosphere. In fact, the temperature obtained using  $OH(7-2)$  band at Kodaikanal showed an increase about 15 minutes after the onset of sub-storm. Similar increase in temperature has also been found by observers [37,38] within a few hours of

sudden commencement of the magnetic storm over very low latitude zones around 200 km altitude. After revealing the unusual heating events, they concluded by arguing that the deposition energy is localized. It seems feasible that the extra-ionospheric current might have helped in increasing the recombination rate which, in turn, produced more atomic oxygen density or in other words, increased the population of  $O(^1S)$ , following the sub-storm.

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